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*Red clover (*Trifolium pratense* L.) draft genome provides a platform for trait improvement*

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Temporal variations in supraglacial debris distribution on Baltoro Glacier, Karakoram between 2001 and 2012

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25 **Abstract**

26 Distribution of supraglacial debris in a glacier system varies spatially and temporally due to
27 differing rates of debris input, transport and deposition. Supraglacial debris distribution
28 governs the thickness of a supraglacial debris layer, an important control on the amount of
29 ablation that occurs under such a debris layer. Characterising supraglacial debris layer
30 thickness on a glacier is therefore key to calculating ablation across a glacier surface. The
31 spatial pattern of debris thickness on Baltoro Glacier has previously been calculated for one
32 discrete point in time (2004) using satellite thermal data and an empirically based
33 relationship between supraglacial debris layer thickness and debris surface temperature
34 identified in the field. Here, the same empirically based relationship was applied to two
35 further datasets (2001, 2012) to calculate debris layer thickness across Baltoro Glacier for
36 three discrete points over an 11-year period (2001, 2004, 2012). Surface velocity and
37 sediment flux were also calculated, as well as debris thickness change between periods.
38 Using these outputs, alongside geomorphological maps of Baltoro Glacier produced for
39 2001, 2004 and 2012, spatiotemporal changes in debris distribution for a sub-decadal
40 timescale were investigated. Sediment flux remained constant throughout the 11-year
41 period. The greatest changes in debris thickness occurred along medial moraines, the
42 locations of mass movement deposition and areas of interaction between tributary glaciers
43 and the main glacier tongue. The study confirms the occurrence of spatiotemporal changes
44 in supraglacial debris layer thickness on sub-decadal timescales, independent of variation
45 in surface velocity. Instead, variation in rates of debris distribution are primarily attributed to
46 frequency and magnitude of mass movement events over decadal timescales, with climate,
47 regional uplift and erosion rates expected to control debris inputs over centurial to millennial
48 timescales. Inclusion of such spatiotemporal variations in debris thickness in distributed
49 surface energy balance models would increase the accuracy of calculated ablation, leading

to a more accurate simulation of glacier mass balance through time, and greater precision in quantification of the response of debris-covered glaciers to climatic change.

Keywords: Karakoram, debris-covered glaciers, supraglacial debris, Baltoro Glacier

1. Introduction

Debris-covered glaciers are commonly found in tectonically-active mountain ranges including the Andes, the Southern Alps of New Zealand and the Himalaya-Karakoram (Kirkbride, 1999; Scherler et al., 2011). High rates of rock uplift and erosion and steep hillslopes in these regions cause large volumes of rock debris to be incorporated into glacier systems, and ultimately form supraglacial debris layers of varying thicknesses and extents (Anderson and Anderson, 2016; Shroder et al., 2000). The presence of a supraglacial debris layer affects ablation of the underlying ice (Evatt et al., 2015; Östrem, 1959), because the debris acts as a thermal barrier between ice and atmosphere, ultimately resulting in a non-linear response of debris-covered glaciers to climatic change (Benn et al., 2012; Scherler et al., 2011). Glaciers in the Himalaya-Karakoram supply water to some of the largest rivers in the world, including the Indus, Brahmaputra and Ganges (Bolch et al., 2012). Consequently, the response of glaciers in the Himalaya-Karakoram to recent and current climatic change will affect the lives of the 1.4 billion people in central Asia who rely on these rivers as their primary water resource (Immerzeel et al., 2010).

Given that the proportion of debris-covered glacier ice area in the Himalaya-Karakoram region is increasing (Deline, 2005; Mihalcea et al., 2006), gaining a full understanding of the influence of debris layers on melt-rates is becoming increasingly pertinent. Typically,

75 supraglacial debris is initially entrained into lateral and medial moraines in the upper
76 reaches of the glacier. As moraines coalesce with increasing distance from their source the
77 debris layer becomes more spatially extensive (Anderson, 2000; Kirkbride and Deline,
78 2013). The thickness of the supraglacial debris layer increases down-glacier and reaches
79 its maximum near the glacier terminus (Anderson, 2000). In areas where supraglacial
80 debris cover extends across the entire glacier surface spatially variable debris distribution
81 results in differential melting and forms an undulating glacier surface topography (Hambrey
82 et al., 2008; Kirkbride and Deline, 2013). Supraglacial debris thickness varies in space and
83 time as a result of differing spatial extents and temporal rates of debris input, transport and
84 exhumation (Rowan et al., 2015). Ablation rates of debris-covered glaciers are therefore
85 also spatially and temporally variable (Benn et al., 2012; Rounce and McKinney, 2014).
86 Studies that consider the response of debris-covered glaciers to climatic change currently
87 do not account for this variability (e.g. Bolch et al., 2012; Scherler et al., 2011; Shea et al.,
88 2015), which increases the uncertainty in estimations of glacier ablation rates, and thus the
89 subsequent predictions of the response of debris-covered glaciers to climatic change.

90

91 The impact of supraglacial debris layers on melt rates is well established (e.g. Östrem,
92 1959); thin debris layers (typically <0.05 m thick, depending on local conditions) enhance
93 ablation by increasing albedo of the glacier surface, while thicker debris layer attenuate
94 melt by insulation of the underlying ice (Mihalcea et al., 2008b; Nicholson and Benn, 2006;
95 Östrem, 1959). Ablation is maximized at an effective debris thickness (commonly 0.01–0.02
96 m), while the critical thickness of debris (ranging from 0.02 to 0.1 m), where ablation under
97 debris-covered ice is equal to that of debris-free ice, is defined by debris properties such as
98 lithology, porosity, grain size distribution, moisture content and surface roughness of the
99 debris layer (Brock et al., 2010; Kayastha et al., 2000). The amount of ablation under a

100 debris layer is also affected by external factors such as the transfer rate of precipitation
101 through a debris layer, glacier surface topography, and the occurrence of suprafluvial
102 networks and associated sediment transport processes, all of which are spatially and
103 temporally variable (Seong et al., 2009).

104

105 Measuring the thickness distribution of a supraglacial debris layer is challenging in the field
106 due to high spatial variability in debris layer thickness over short distances, difficulties in
107 excavating such debris layers (Mayer et al., 2006), and an inability to capture such
108 variability with point data. Early work put forward the idea of using thermal characteristics of
109 supraglacial debris to define its extent from satellite data (Ranzi et al., 2004). Subsequent
110 projects developed the use of such thermal satellite data to estimate debris thickness for
111 entire glacier surfaces: a glacier-specific relationship between surface temperature and
112 debris thickness is identified using field point data, which is subsequently applied to
113 satellite-derived thermal data of the entire glacier area (e.g. Foster et al., 2012; Rounce and
114 McKinney, 2014; Mihalcea et al., 2008a; Mihalcea et al., 2008b; Soncini et al., 2016). These
115 maps have advanced understanding of spatial variability in debris thickness, but usually
116 only represent a discrete point in time. Minora et al. (2015) enabled the observation of
117 temporal changes in debris thickness by producing a second debris thickness map of
118 Baltoro Glacier for 2011, in addition to the one produced by Mihalcea et al. (2008b) for
119 2004. However, Minora et al. (2015) did not explore the extent of debris thickness change
120 between the two periods. Consequently, little is known about the rate at which changes in
121 supraglacial debris layer thickness occur, an essential parameter for understanding the
122 transport of debris by ice flow and the localised redistribution of the debris over a glacier
123 surface, which can be used to validate precise numerical modelling of the dynamics of

124 debris-covered glaciers through time (e.g. Anderson and Anderson, 2016; Rowan et al.,
125 2015).

126

127 In this study, we investigated supraglacial debris on Baltoro Glacier in Pakistan to: (1)
128 identify the spatio-temporal variation in supraglacial debris distribution on Baltoro Glacier
129 between 2001 and 2012; (2) consider some of the processes that control these variations
130 in debris distribution using surface velocity and geomorphological mapping; and (3)
131 calculate annual rates of debris thickness change and sediment flux on Baltoro Glacier
132 using debris thickness and surface velocity maps. We subsequently comment on how such
133 calculations can be used in numerical models for glaciers.

134

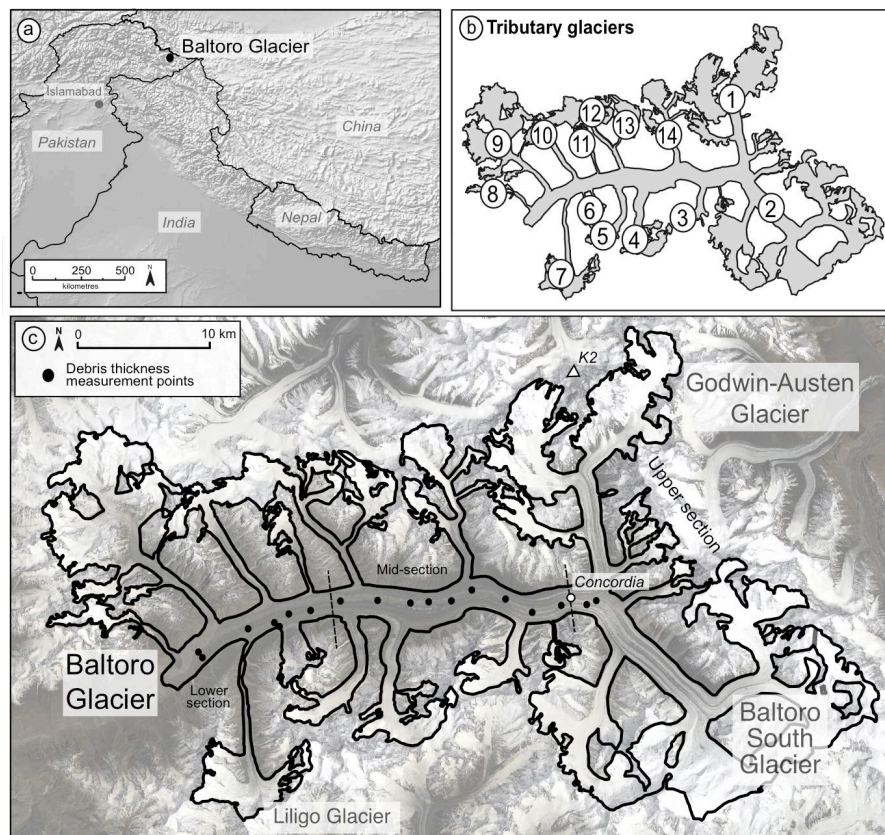
135 **2. Study area**

136 Baltoro Glacier is located in the eastern Karakoram mountain range in northern Pakistan
137 (35°35' N, 76°04' E; Figure 1). The glacier is 62 km long and flows from near the peak of K2
138 (8611 m a.s.l.) to an altitude of 3410 m a.s.l. (Mayer et al., 2006; Mihalcea et al., 2008b)
139 (Figure 1a). A number of tributary glaciers feed Baltoro Glacier (Figure 1b), including
140 Baltoro South and Godwin-Austen Glaciers, which converge to form the main Baltoro
141 Glacier tongue at Concordia (4600 m). The surface velocity of Baltoro Glacier varies along
142 its length, with a maximum surface velocity of $\sim 200 \text{ m a}^{-1}$ below Concordia, decreasing to
143 less than 15 m a^{-1} close to the glacier terminus (Copland et al., 2009; Quincey et al., 2009).
144 Surface velocity was observed to increase in 2005 (Quincey et al., 2009), attributed to an
145 abundance of meltwater being routed to the bed and thus reducing basal friction.

146

147 The ablation area of Baltoro Glacier is almost entirely debris covered. Up-glacier of
148 Concordia, supraglacial debris is predominantly entrained into medial and lateral moraines

149 that punctuate the clean ice surface, with a lesser contribution of mass movement deposits
 150 along the glacier margins (Mihalcea et al., 2006). The debris layer is thinnest (0.01–0.15 m)
 151 in the upper ablation area of the glacier and exceeds 1 m at the glacier terminus (Mihalcea
 152 et al., 2008b). Supraglacial debris units have differing lithologies across the debris-covered
 153 glacier surface, which include granite, schist, gneiss and metasediments (Gibson et al.,
 154 2016).



155
 156 **Figure 1.** (a) Baltoro Glacier in a regional context; (b) the tributary glaciers of Baltoro
 157 Glacier (numbered) and (c) Baltoro Glacier and its tributary glaciers.

158

159 3. Methods

160 3.1. Debris thickness

161 Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) thermal
162 data were used to derive debris thickness on Baltoro Glacier for three discrete periods in
163 time; 2001, 2004 and 2012 (Table 1). The 2004 dataset was the same as that used by
164 Mihalcea et al. (2008b) for production of their 2004 debris thickness map of Baltoro Glacier.
165 The 2001 and 2012 data sets were chosen due to their low cloud cover, resulting in minimal
166 glacier area being obscured. ASTER imagery was downloaded from NASA's Earth
167 Observing System Data and Information System (<http://reverb.echo.nasa.gov>) as a Level 2
168 surface kinetic temperature product (AST_08). Level 2 surface kinetic temperature data are
169 comprised of mean surface temperature calculated from thermal bands 11–15. Prior to
170 delivery, surface kinetic temperature data are atmospherically corrected and converted from
171 top-of-atmosphere temperature to surface temperature. ASTER thermal data have a spatial
172 resolution of 90 m and temperature resolution of 0.5 K (Abrams and Ramachandran, 2002).
173 The 2001 and 2012 ASTER datasets were both acquired in August within 15 days of the
174 original 2004 dataset, therefore allowing for seasonal variation in debris surface to be
175 minimised as much as possible when comparing subsequent outputs. All outputs were co-
176 registered to within a pixel through manual placement of 50 tie points between each image
177 pair prior to calculation of debris thickness, to avoid any spatial mismatch between the input
178 layers. The images were also orthorectified using the rigorous orthorectification tool in
179 ENVI (v. 5.0) and the ASTER digital elevation model (2011) at 30 m resolution, which
180 corrects for the effect of sensor tilt and terrain, and produced an RMSE of 5.82 m. Debris
181 thickness was derived using the methods detailed in Mihalcea et al. (2008b). Equation 1
182 was applied to the same satellite image used by Mihalcea et al. (2008b) to yield debris
183 thickness for 2004:

184

$$185 \quad DT = \exp (0.0192 T_s - 58.7174) \quad (1)$$

186

187 Where DT is debris thickness and T_s is surface temperature. The same method was then
 188 applied to the 2001 and 2012 ASTER data for Baltoro Glacier to yield a time-series of
 189 debris thickness maps.

190

191

192

193 **Table 1.** Satellite ID, acquisition date and time and mean debris thickness for ASTER
 194 datasets used for calculating debris thickness (grey boxes) and surface velocity.

195

Satellite data I.D.	Acquisition date and time	Mean debris thickness (m)
AST_08_00308292001060003_20140108123858_15605	29/08/2001 06:00	0.14
AST_08_00310032002055404_20151109052624_814	03/09/2002 05:54	
AST_08_00308142004054614_20151109052424_30691	14/08/2004 05:46	0.21
AST_08_00310122008054700_20151109052644_888	12/09/2008 05:47	
AST_08_00305052011055248_20151109052354_30515	05/09/2011 05:52	
AST_08_00308202012054630_20151109052624_816	20/08/2012 05:46	0.45

196

197

198

199 Debris thickness change was calculated between the 2001–2004 and 2004–2012 debris
 200 thickness maps. In both cases the earlier debris thickness map was subtracted from the

201 later map to yield debris thickness change for each time period, and then divided by the
202 number of years between the two maps to calculate mean annual debris thickness change.

203

204 Uncertainty in the calculated debris thickness was estimated for the 2012 debris thickness
205 map using field debris thickness measurements collected in 2013 (Figure 1c). Mean annual
206 debris thickness change calculated from the 2004–2012 debris change map (0.03 m a^{-1})
207 was added to the 2012 debris thickness to provide projected debris thickness for 2013. To
208 calculate uncertainty the 17 field-derived debris thickness point measurements were
209 compared to debris thicknesses from the corresponding pixels in the projected 2013 debris
210 thickness map (Table 2). Mean variation in debris thickness between 2013 field data and
211 projected 2013 debris thickness was 0.090 m, 0.064 m above the uncertainty calculated for
212 the 2004 debris thickness map by Mihalcea et al. (2008b) of 0.026 m. Consequently, in this
213 study uncertainties for the debris thickness maps were estimated as 0.026 m for 2004 and
214 0.090 m for 2012. Uncertainty for the 2001 debris thickness map could not be calculated
215 due to a lack of field data collected prior to 2004. Additional parameters such as moisture
216 content in and thermal inertia of the debris layer may have also affected estimations of
217 supraglacial debris layer thickness calculated using Mihalcea et al.'s (2008b) method, but the
218 low uncertainty values calculated here suggest they have minimal effect on the outputs
219 presented.

220

221 Due to a lack of field data in debris layers with a thickness greater than 0.5 m uncertainty
222 values calculated here are only applicable for debris layers with a thickness $\leq 0.5 \text{ m}$. Above
223 0.5 m debris surface temperature is considered independent of debris layer thickness
224 (Nicholson and Benn, 2006). Consequently, analysis of these debris thickness maps is

225 focused on areas of the glacier where debris thickness is ≤ 0.5 m, and analysis of debris
 226 thickness maps are presented alongside geomorphological evidence for justification.

227

228

229 **Table 2.** Comparison of field point debris thickness data to corresponding pixel value (plus
 230 one year's annual rate of debris thickness change), used to calculate error between the
 231 2012 satellite-derived debris thickness map and field data.

Point I.D.	2013 <i>in situ</i> debris thickness (m)	2013 satellite- derived debris thickness (m)	Difference (m)
1	0.02	0.01	0.01
2	0.00	0.07	0.07
3	0.09	0.23	0.14
4	0.13	0.15	0.02
5	0.01	0.04	0.03
6	0.06	0.03	0.03
7	0.04	0.02	0.02
8	0.05	0.14	0.09
9	0.075	0.02	0.05
10	0.04	0.23	0.19
11	0.12	0.39	0.27
12	0.07	0.18	0.11
13	0.17	0.06	0.11
14	0.04	0.14	0.10
15	0.26	0.28	0.02
16	0.1	0.01	0.09
17	0.43	1.16	0.73
Mean difference:			0.09
Standard deviation:			0.50

232

233

234 **3.2. *Glacier dynamics and surface morphology***

235 3.2.1. Surface velocity analysis

236 Glacier surface velocity analysis was undertaken in ENVI (v.5.0) using the feature tracking
237 plugin tool Cosi-Corr (Leprince et al., 2007). Cosi-Corr is a Fourier-based image correlation
238 tool that offers sub-pixel accuracy for the measurement of horizontal offsets (Scherler et al.,
239 2011). ASTER Band 3N data (Visible Near Infrared, Wavelength: 0.760–0.860 nm,
240 resolution: 15 m) were used for feature tracking. Image pairs used were acquired in 2001
241 and 2002, 2003 and 2004, and 2011 and 2012 (Table 1), and were co-registered to sub-
242 pixel level prior to calculation of surface displacement. All results were converted to annual
243 displacements for comparison. A variable window size between 128 and 64 pixels and a
244 step size of one pixel was used for all velocity outputs, and absolute surface velocity
245 derived from north-south and east-west velocity fields. North-south and east-west velocity
246 fields were used for identification of direction of maximum surface velocity in the calculation
247 of sediment flux. Velocity outputs were masked using a velocity threshold of 200 m a^{-1} to
248 exclude erroneous results in ENVI (v. 5.0), and clipped to the extent of a manually-
249 improved Baltoro Glacier outline based on the Randolph Glacier Inventory outline (v. 5.0;
250 Arendt et al., 2012), used in Gibson et al. (2016) in ArcMap (v.10.1). Pixels with erroneous
251 surface velocity values (less than zero or above 200 m a^{-1} , or pixels with substantially
252 different velocity values to the surrounding pixels) were masked from the final surface
253 velocity maps.

254

255 3.2.2. *Geomorphological mapping*

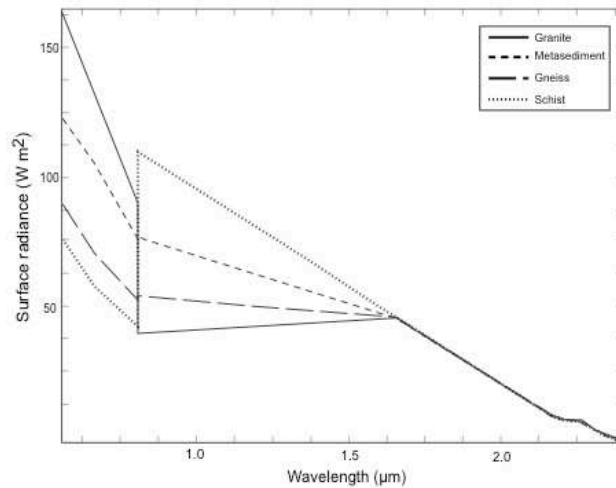
256 Geomorphological features on the surface of Baltoro Glacier, including debris units, mass
257 movement deposits, supraglacial water bodies and crevasses, were mapped using the

optical bands (15 m resolution) of the same three time-separated and orthorectified ASTER data sets used for deriving debris thickness (August 2001, 2004 and 2012). ASTER images were orthorectified using the ASTER digital elevation model at 30 m resolution. Additionally, a Landsat 7 Enhanced Thematic Mapper (ETM+) image, acquired in August 2001 at 30 m resolution, was used to map regions of the glacier covered in cloud in the ASTER August 2001 imagery. All satellite datasets used were co-registered prior to mapping. Features were mapped using a false-colour composite (Bands 3N, 2, 1) and were manually digitised in ESRI ArcGIS (v. 10.1). Debris units were classified using their differing spectral reflectance profiles (Figure 2; Lillesand et al., 2014). The debris units were then traced up-glacier to their source area and lithology identified using the regional geological map produced by Searle et al. (2010). Spectra from 200 pixels were then compared to spectra from the USGS spectral library in ENVI to confirm correct classification. 91% of sampled pixels were correctly classified based on these independent data. Mass movement deposits were identified by the presence of two features: a scar, identified as an elongate feature on the valley side which differed in colour to the surrounding valley wall, suggesting erosion and loss of vegetation had occurred, and an associated lobate debris fan deposit on or near the glacier surface. A Normal Difference Water Index (NDWI) was used to identify pixels containing supraglacial water, calculated from ASTER bands 3 (Near Infrared; NIR) and 4 (Shortwave Infrared: SWIR) after (Gao, 1996):

$$NDWI = (NIR (Band 3) - SWIR (Band 4)) / (NIR (Band 3) + SWIR (Band 4)) \quad (2)$$

The classification of water was verified through manual comparison of 100 randomly selected features classified as water in the 2011 ASTER imagery with high resolution (2.5 m) Quickbird imagery from 2011 for the same locations. All 100 features were identified as

283 water in both images, and so were assumed to be correctly classified. The area of debris
284 units and supraglacial water bodies were derived using the geometry calculator in ArcGIS
285 (v.10.1).



286
287 **Figure 2:** Spectral reflectance profiles for the different lithology types identified on Baltoro
288 Glacier.

289
290 **3.3. Sediment flux**

291 Supraglacial sediment flux across the glacier surface was calculated using derived debris
292 thickness and surface velocity data, following the method developed by Heimsath and
293 McGlynn (2008) to determine headwall retreat rate on Milarepa's Glacier in Nepal.
294 Heimsath and McGlynn (2008) measured debris thickness and surface velocity along one
295 transect near the glacier headwall, then calculated cross-sectional area of the debris using
296 the debris thickness transect, and multiplied the cross-sectional area by surface velocity,
297 calculating one-dimensional sediment flux. Here, we calculated supraglacial sediment flux
298 for each pixel by multiplying debris thickness by the pixel width at right angles to the
299 direction of maximum surface velocity to give supraglacial debris layer cross-sectional area,

300 and then multiplied cross-sectional area by surface velocity for the same pixel. As surface
301 velocity and supraglacial debris thickness were used to calculate sediment flux these
302 results only represent debris transported supraglacially. The resulting sediment flux maps
303 were normalized to annual datasets to obtain comparable sediment flux values, and were
304 masked using the same masks applied to the surface velocity and debris thickness maps to
305 exclude pixels with erroneous results and cloud cover.

306

307 **4. Results**

308 **4.1. Debris thickness**

309 A similar pattern of debris thickness distribution was present in 2001, 2004 and 2012
310 (Figure 3). In the upper section of the glacier above and around Concordia, debris was
311 distributed in alternating bands of thicker debris (around 0.2–0.3 m thick) and thin, sparsely-
312 distributed or non-existent debris layers (≤ 0.02 m), in a longitudinal pattern parallel to ice
313 flow. Thicker bands of debris originated from the glacier margin, primarily at confluences
314 between tributary glaciers and the main glacier tongue, which were interpreted to be medial
315 moraines. In the glacier mid-section, debris coverage became increasingly spatially
316 extensive with decreasing distance from the glacier tongue, and a general thickening of
317 debris towards the glacier terminus occurred. No build up of debris, such as that expected
318 where a terminal moraine is present, was observed at the glacier terminus from satellite
319 data, confirming the absence of such a feature previously observed in the field by Desio
320 (1954). Debris covered the entire glacier surface in the lower section of the glacier and was
321 predominantly > 0.5 m thick.

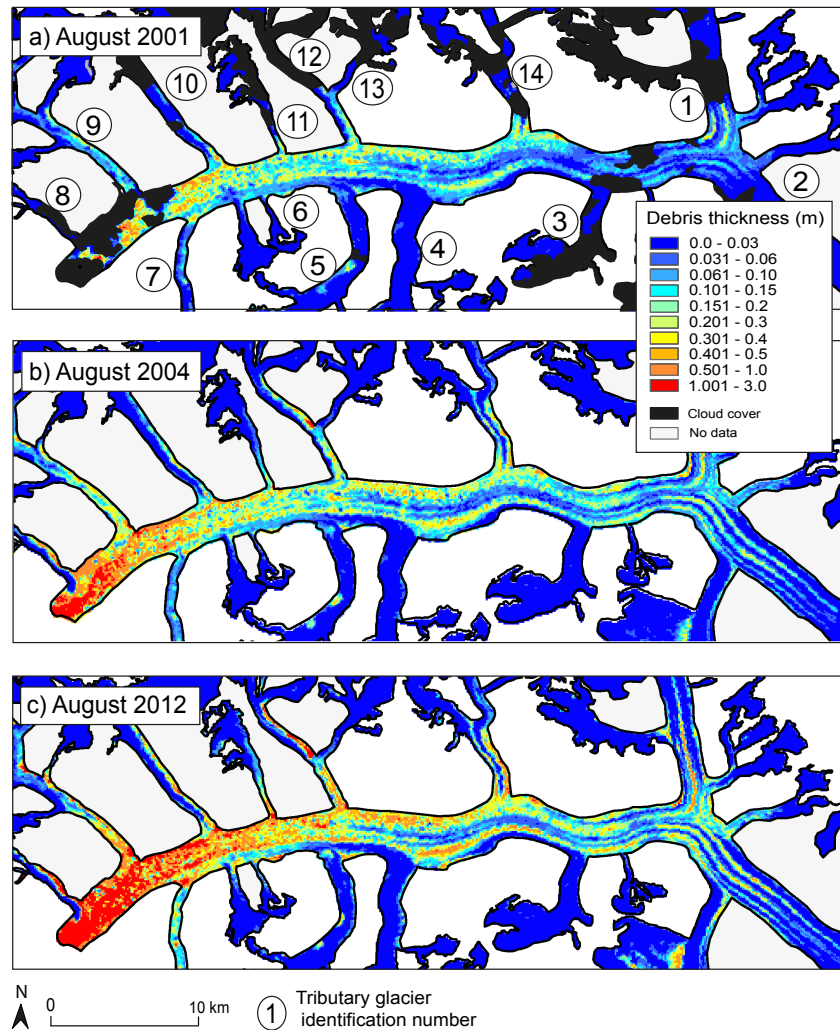


Figure 3: Debris thickness maps of Baltoro Glacier for: (a) August 2001; (b) August 2004; (c) August 2012.

The broad, glacier-wide pattern of debris distribution displayed minimal change between 2001 and 2012, suggesting that a pattern of debris input and transport was already established across the glacier and persisted over the study period. However, the thickness of the debris layers across the glacier varied over the 11-year study period. Cloud cover in 2001 restricted comparison between 2001 and 2004 in the glacier's lower section, but

331 thickening of the medial moraines in the glacier upper-section of the order of around 0.1m
332 was seen during this 3-year time period. A general trend of increasing debris thickness in
333 the glacier mid-section was seen between 2001 and 2012, with a mean debris layer
334 thickness in the glacier mid-section of ~0.28 m in 2001, ~0.34 m in 2004 and ~0.41 m in
335 2012. Debris thickness was most variable in the glacier lower section between 2004 and
336 2012, with a mean debris layer thickness of ~0.71 m in 2004 and ~1.5 m in 2012, and an
337 apparent thickening of debris at the terminus, although further field data would be needed
338 to confirm these mean debris thicknesses due to the independence of debris surface
339 temperature with debris layer thickness above 0.5 m. Increasing debris thickness in the
340 lower and mid sections suggests a progressive backing up of debris through time causing
341 the area of thickest debris to increase up-glacier from the terminus.

342

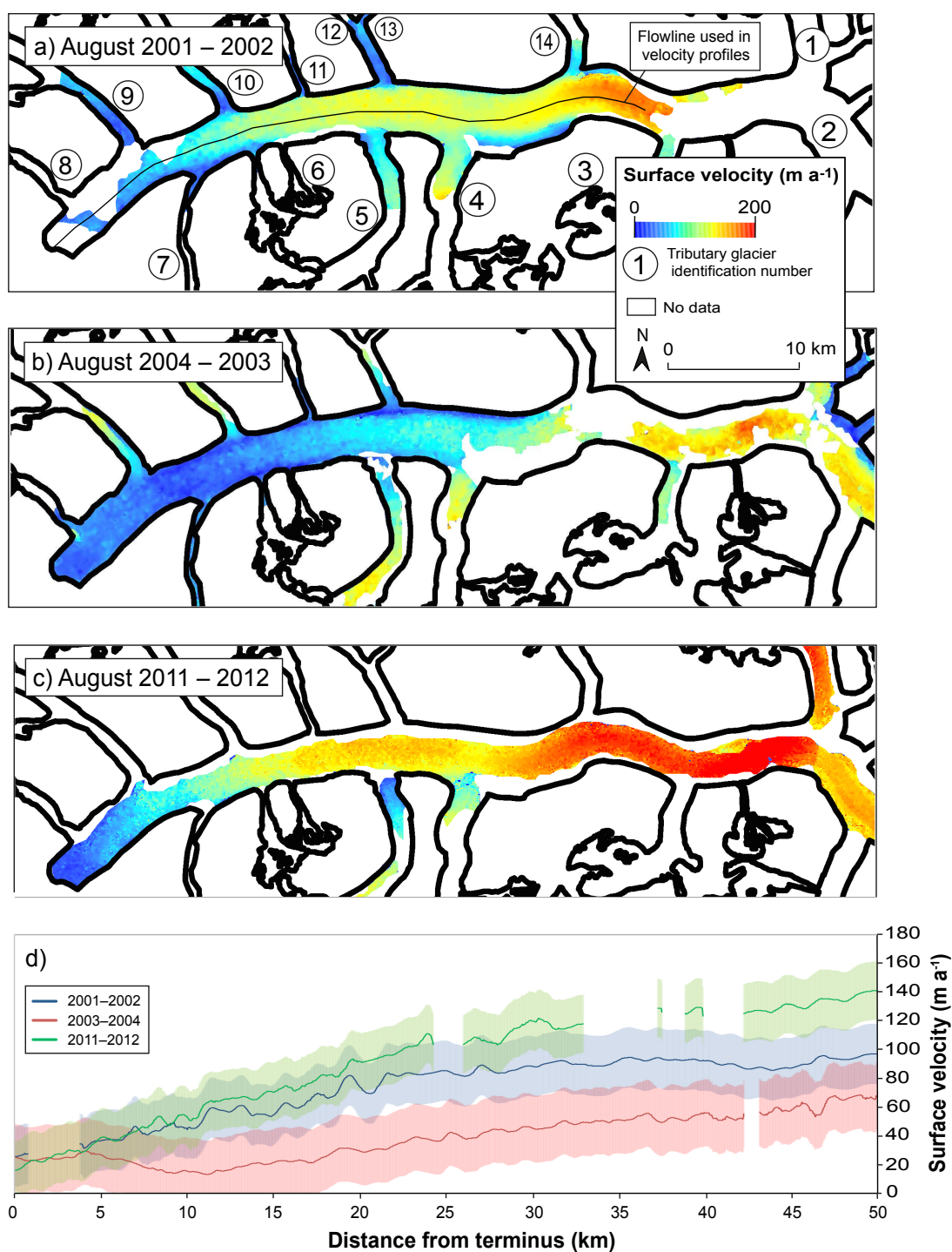
343 In 2004 a sharp boundary between debris layer thicknesses was observed running
344 longitudinally from the glacier terminus to the location at which Trango Glacier (Tributary
345 Glacier 9) joins the main glacier tongue (Figures 1b; 3). South of the boundary debris
346 thickness was above 0.5 m thick, whilst north of the boundary debris layer thickness was
347 less than 0.5 m thick. The debris thickness boundary correlates with the boundary between
348 a granite debris unit originating on Trango Glacier and gneiss debris units of the main
349 glacier tongue, presumed to also be the boundary between the main glacier flow units and
350 Trango Glacier flow unit.

351

352 **4.2. Glacier surface velocity**

353 A general trend of highest velocity at Concordia, where Baltoro South and Godwin-Austen
354 Glacier converge, and subsequently decreasing surface velocity down-glacier of Concordia
355 towards the terminus was observed at all time periods, with very low (less than 20 m a^{-1}) to

no glacier flow near the terminus (Figure 4). Variations in surface velocity occurred between 2001 and 2012, with an average decrease in surface velocity of around 50 m a^{-1} along the longitudinal profile of the glacier (Figure 4a) between 2001 and 2004, followed by an increase on the same order of magnitude between 2004 and 2012 (Figure 4d). Higher surface velocities were observed at Concordia where the Godwin-Austen and Baltoro South Glaciers join, and subtle velocity increases at some but not all tributary glacier confluences were also noted (e.g. Yermanendu and Mandu glaciers; Tributary glaciers 4 and 5, respectively). In 2012 glacier surface velocity was lowest ($\sim 0\text{--}20 \text{ m a}^{-1}$) in the northwest region of the terminus, a triangular shaped area which extended from the glacier terminus and pinched out at around 5 km up-glacier of the terminus and downstream of Trango Glacier. However, in 2004 no such pattern was evident and a patchy distribution of velocity between 0 and 50 m a^{-1} across the glacier width for around 10 km up-glacier of the terminus occurred.



369

370 **Figure 4:** Surface velocity maps in m a^{-1} for Baltoro Glacier for: (a) 2001, (b) 2004 and (c)
 371 2012, and (d) surface velocity profiles along the centre line of the main glacier tongue with
 372 uncertainty values of each velocity line displayed with shaded regions.

373 **4.3. Geomorphological features**

374 Supraglacial debris lithology was identified through comparison of ASTER pixel spectra with
375 spectra of lithologies from the USGS spectral library and with reference to the geology map
376 produced by Searle et al. (2010). The supraglacial debris on Baltoro Glacier was dominated
377 by gneiss (~51–53 % of the debris-covered glacier area), whilst ~47–49 % was composed
378 of granite (~27 %), schist (~12 %) and a small proportion of metasediment (~6 %) (Figure
379 5, Table 3). Across the main glacier tongue, negligible change in debris unit boundaries
380 occurred between 2001 and 2012 and change in percentage cover of debris units was
381 attributed to errors produced by manual digitisations (Gibson et al., 2016; Table 3).
382 However, small scale variations in debris distribution did occur on tributary glaciers between
383 2001 and 2012, which have been attributed to these glaciers being in various periods of
384 instability, possibly related to surge phases,, and input of debris material from surrounding
385 valley walls through rock- and snow avalanches (Gibson et al., 2016). For example,
386 patches of thicker debris on Mandu Glacier (Tributary Glacier 5) can be tracked down-
387 glacier between 2001 and 2012 in geomorphological maps (Figure 6d) and debris thickness
388 maps (Figure 6e), with debris initially deposited on the glacier by a mass movement event
389 and then transported as a bulk volume.

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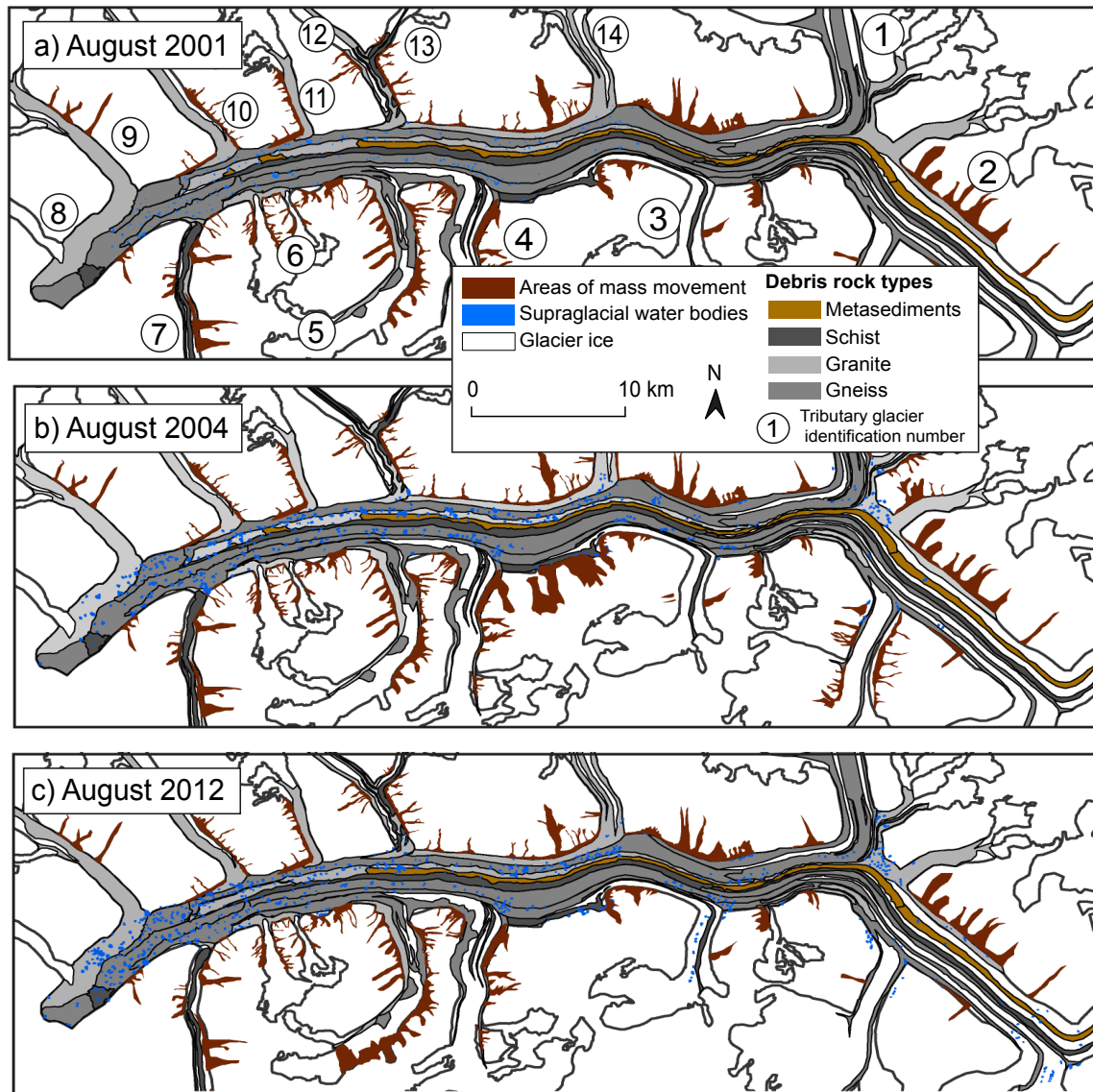
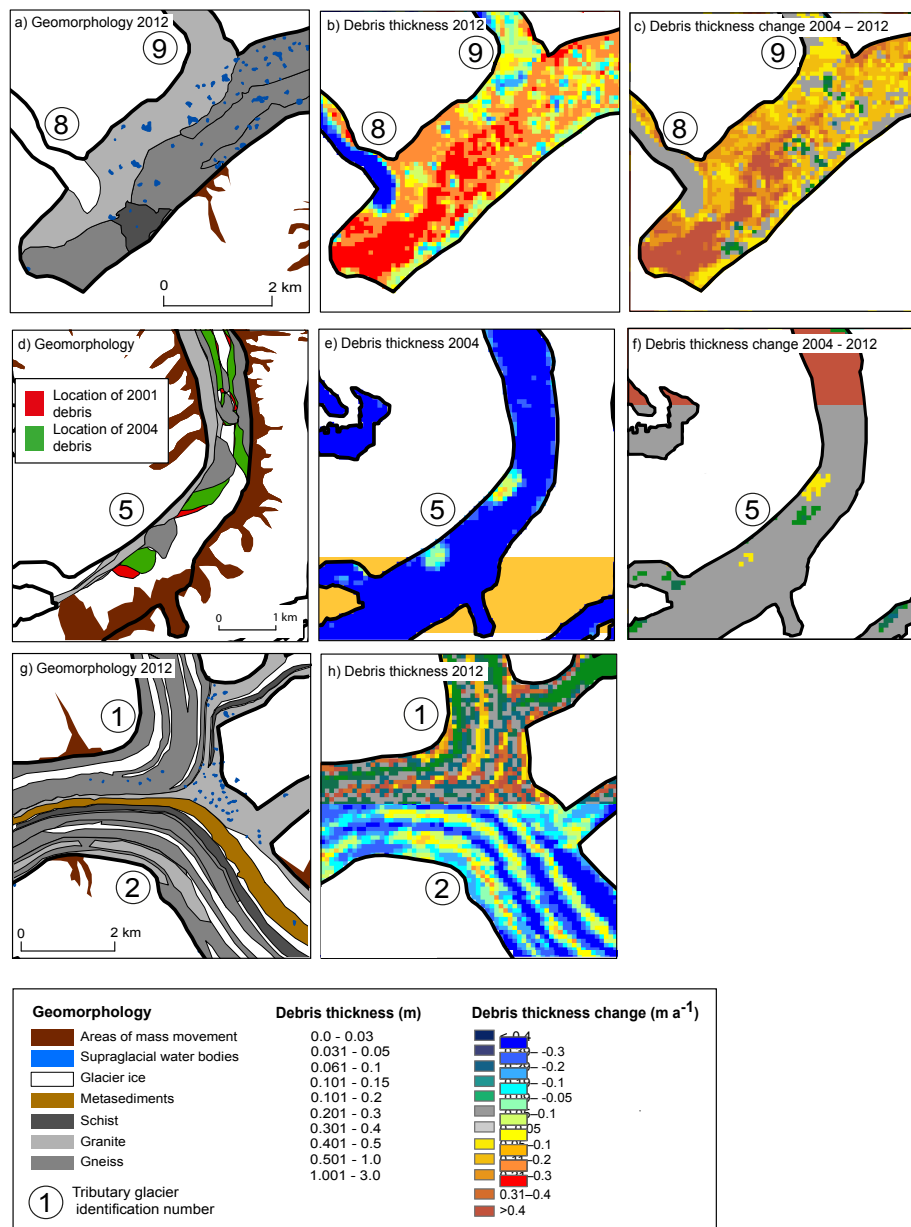


Figure 5: The surface geomorphology of Baltoro Glacier in (a) August 2001, (b) August 2004 and (c) August 2012.

Table 3. Total area of each debris unit type, based on lithology, for 2001, 2004 and 2012, and the percentage of each debris type as a proportion of the total debris cover for Baltoro Glacier and its tributary glaciers (Gibson et al., 2016). Variability in total debris area is attributed uncertainty produced by manual digitisation.

Year	2001	2001	2004	2004	2012	2012
Debris type	Area (km ²)	% of total debris	Area (km ²)	% of total debris	Area (km ²)	% of total debris
Gneiss	81.48	52.9	79.83	51.2	79.48	52.9
Metasediment	8.60	5.6	11.41	7.3	8.28	5.5
Schist	17.76	11.5	18.54	11.9	17.71	11.8
Granite	46.29	30.0	46.20	29.6	44.71	29.8
Total debris (km²)	154.13		155.97		150.17	



415 **Figure 6:** Comparison of: (a) geomorphology, (b) debris thickness and, (c) debris thickness
 416 change at the terminus of Baltoro Glacier, showing a distinct boundary between debris
 417 thickness values and debris units of different lithologies; (d) geomorphology, with previous
 418 positions on areas of supraglacial debris from 2001 (red) and 2004 (green) displayed, (e)
 419 debris thickness and (f) debris thickness change on Mandu Glacier (Tributary Glacier 5)
 420 showing the down-glacier movement of debris pockets through time; (g) geomorphology
 421 and (h) debris thickness of the confluence area between Godwin-Austen Glacier and
 422 Baltoro South Glacier in 2012, showing an area of thick debris up-glacier of where the
 423 tributary glaciers join and change direction to form the main glacier tongue.

Supraglacial water bodies occurred most frequently in the lower to lower-mid-sections of the glacier between 2001 and 2012, and in areas of relatively thick debris, such as east of the confluence between Goodwin-Austen and Baltoro South Glaciers (Figure 6g and 6h). However, at all time points an absence of supraglacial water bodies was present in the gneiss debris unit at the terminus of the main glacier tongue and around the terminus of Tributary Glacier 8 (unnamed). The number of supraglacial water bodies increased by 336 over the study period, from 234 in 2001 to 570 in 2012 (Table 4). Total area of supraglacial water bodies increased by almost 400 % during the same period. Temporally, the greatest change in water body number and area occurred between 2001 and 2004, whilst spatially the greatest increase in water body number occurred in the lower mid-section and east of Concordia at up-glacier margin of the confluence between Godwin-Austen and Baltoro South Glacier.

Table 4. Supraglacial water area and number on Baltoro Glacier in 2001, 2004 and 2012 (Gibson et al., 2016).

	2001	2004	2012
Number of water bodies	234	404	570
Area (km²)	0.66	1.79	2.04

4.4. Annual debris thickness change

Mean annual debris thickness change (Figure 7) showed areas of debris thickness increase predominantly occurred in the lower section of the glacier and along medial moraines, and were of the order of 0.05 to 0.3 m a⁻¹, greater than uncertainty values for debris thickness

445 maps. In areas of decreasing debris thickness a reduction in thickness of the order of 0.05–
446 0.09 m a⁻¹ was observed, with most change occurring between 2001 and 2004, although
447 these areas of decrease were lower than the uncertainty values associated with the debris
448 thickness maps. Such areas of decreasing debris thickness occurred on the northern
449 margin of the main glacier tongue and parallel to debris layer thickening of medial
450 moraines. Debris thickening occurred at a similar rate and pattern in the lower section of the
451 glacier between the two periods, with the greatest increase along the boundary between the
452 main glacier tongue and Trango Glacier (Section 4.1). During both periods, increase in
453 debris thickness was primarily along the moraine crests in the mid-section of the glacier,
454 with more extensive increases between 2001 and 2004, extending to the glacier upper-
455 section. Debris thickness change on tributary glaciers was of the order of ± 0.05 m a⁻¹, with
456 specific areas of debris change apparent, including deposits on Mandu Glacier, considered
457 to have been derived from mass movement events which moved down-glacier through
458 time, revealed through a loss of thickness in their previous position and an increasing
459 debris thickness in the current position (Figure 6f).

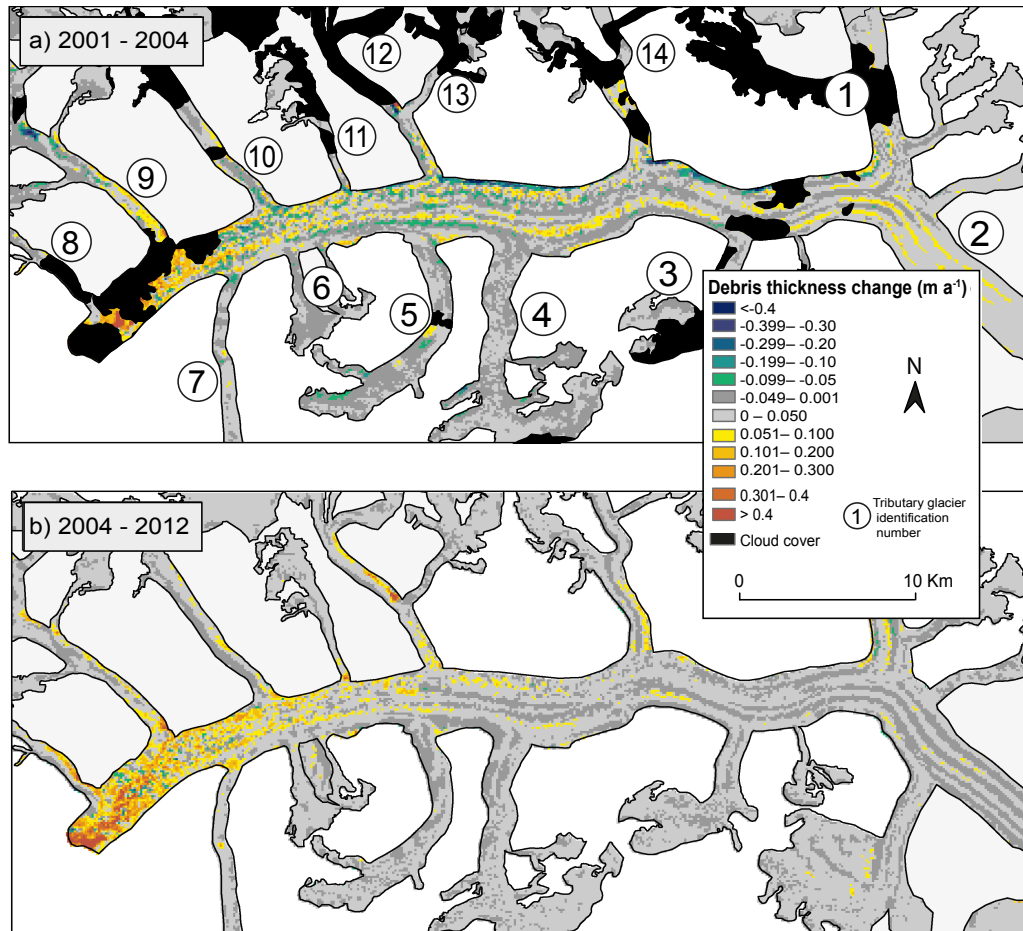


Figure 7: Annual debris thickness change, calculated by subtracting the earlier debris thickness map from the later, and then divided by the number of years between the two maps, for (a) 2001 – 2004 and (b) 2004 – 2012.

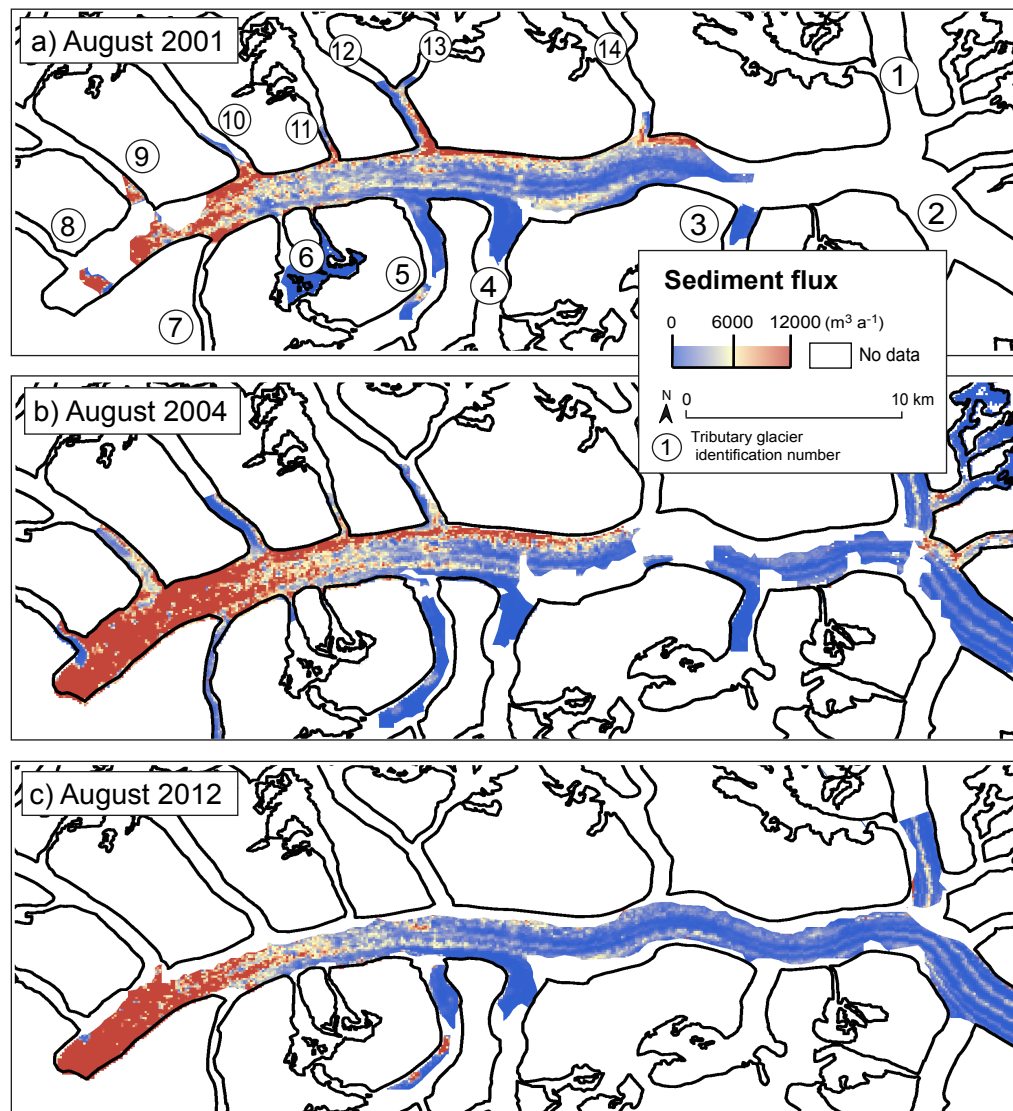
4.5. Annual sediment flux

Supraglacial sediment flux (Figure 8) showed a similar spatial distribution for all points in time, with the highest sediment flux (between 11000 and 12000 m³ a⁻¹) in the lower section of the glacier and along the northern glacier margin in the glacier mid-section. Areas of higher sediment flux (>9000 m³ a⁻¹) were also found at the confluence of tributary glaciers and the main glacier tongue, such as east of Concordia (2003-2004), Yermanendu Glacier (Tributary Glacier 4; 2001-2002) and Tributary Glacier 6 (unnamed; 2001-2002, 2003-

472 2004). For a large proportion of the mid- and upper sections of the glacier, sediment flux
473 was generally less than $1000 \text{ m}^3 \text{ a}^{-1}$, with some areas of relatively higher sediment flux
474 along moraine features ($4000\text{--}6000 \text{ m}^3 \text{ a}^{-1}$).

475

476 A general pattern of increasing sediment flux was seen between 2001–2002 and 2011–
477 2012 along medial moraines in the lower section of the glacier, with an increase in sediment
478 flux on the order of between $5000\text{--}6000 \text{ m}^3 \text{ a}^{-1}$ between 2001 and 2002 to $6000\text{--}8000 \text{ m}^3 \text{ a}^{-1}$
479 1 between 2001 and 2012. In the upper-mid and upper-sections of the glacier these medial
480 moraines had a constant sediment flux of around $6000 \text{ m}^3 \text{ a}^{-1}$. Although the sediment flux
481 maps do not extend to the initiation point of many of the medial moraines where debris is
482 introduced into the upper glacier system, consistency in sediment flux along moraine
483 features suggest input from valley wall erosion and entrainment was stable over the sub-
484 decadal period. In the 2001–2002 and 2003–2004 sediment flux maps pockets of sediment
485 flux less than $1000 \text{ m}^3 \text{ a}^{-1}$ in the lower section of the glacier corresponded to the location of
486 supraglacial water bodies.



487 **Figure 8:** Sediment flux; debris cross-sectional area for each pixel multiplied by surface
 488 velocity, for (a) 2001–2002, (b) 2003–2004 and (c) 2011–2012.

489

490

491 **5. Discussion**

492 **5.1. Spatiotemporal change in supraglacial debris distribution**

493 A debris distribution common to the majority of debris-covered glaciers is evident on the
 494 surface of Baltoro Glacier throughout the study period, with the thickest debris occurring
 495 near the terminus and along moraine crests, and an increasingly thick debris layer towards

496 the terminus (e.g. Figure 9; Fushimi et al., 1980; Kirkbride and Warren, 1999; Mihalcea et
497 al., 2006b; Zhang et al., 2011). A progressive increase in the area covered by debris
498 through time would be expected due to debris being constantly transported to the glacier
499 terminus; such a pattern is observed on Baltoro Glacier between 2001 and 2012, and
500 combined with continued glacier flow would result in a build-up of debris in the lower
501 sections (Kirkbride and Warren, 1999), particularly where there is no efficient sediment
502 evacuation down-valley. A mean increase in debris thickness of between 0.05 and 0.10 m
503 across the glacier surface occurred during the study period. Where the debris layer is below
504 0.5 m, the thickness at which ablation of underlying ice is most variable with debris
505 thickness, the rate of debris thickness change identified here could lead to areas of the
506 debris layer evolving from a thickness that enhances melt to one that insulates it over
507 relatively short timescales (e.g. several years). The rapidity of such changes could render
508 debris thickness maps previously published to be inapplicable for any year other than the
509 one in which debris surface temperature data were collected (e.g. Mihalcea et al., 2008),
510 although such maps would still be important for observing historical debris distribution.

511

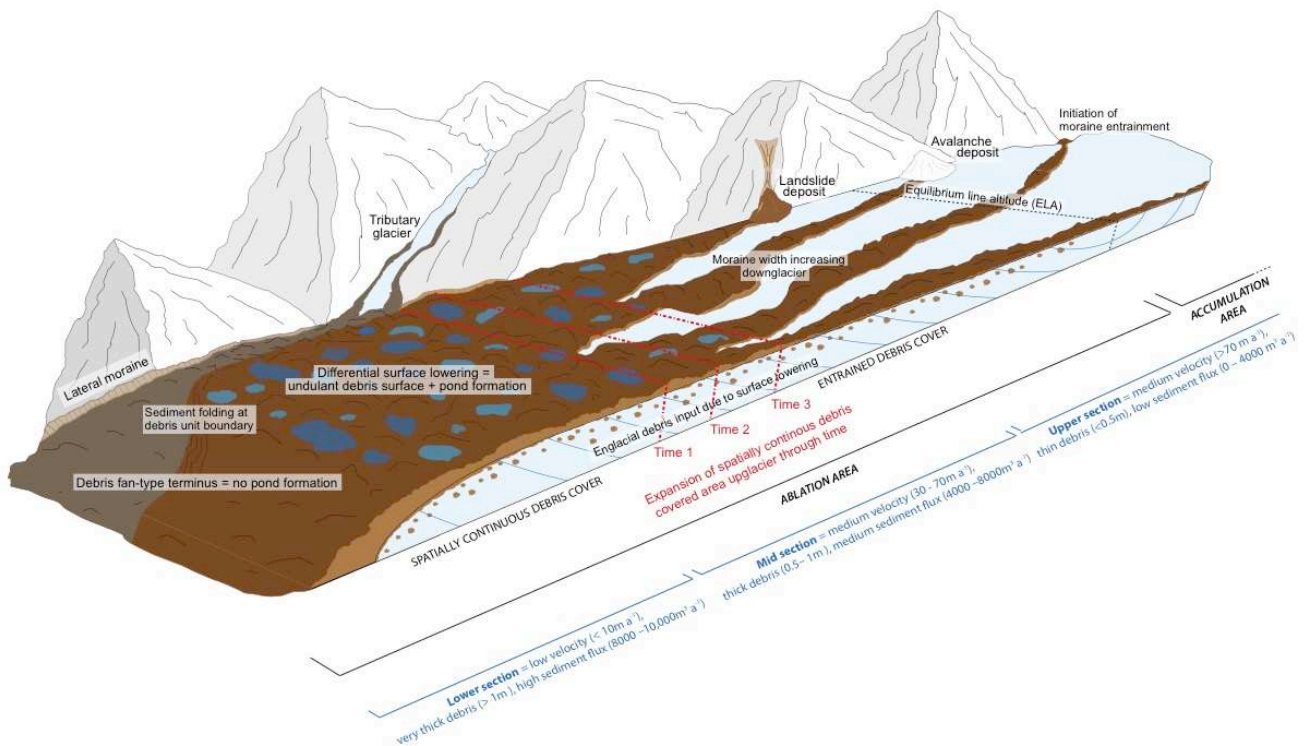
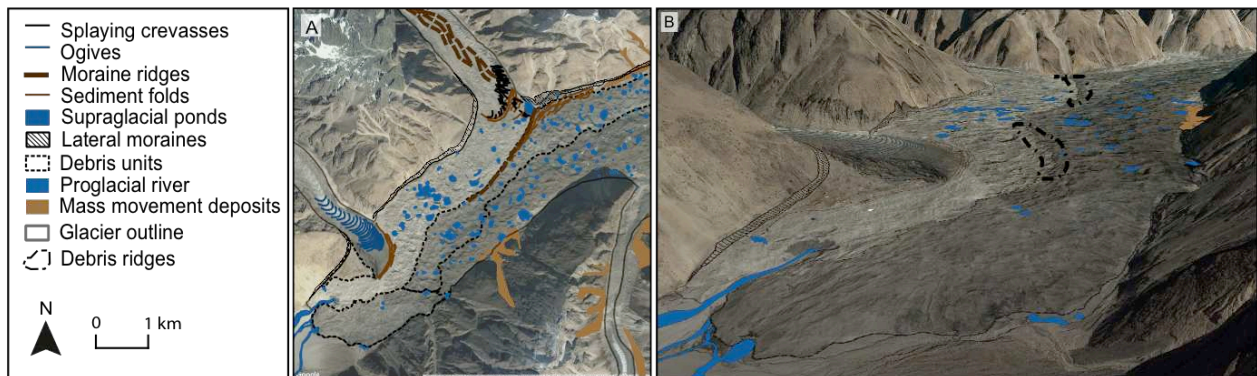


Figure 9: Schematic diagram of a debris-covered glacier system with input, transport and depositional processes alongside glacier dynamics for each section of the glacier, and the change in debris-covered area through time (T_1 – T_3).

Debris thickness change in the lower- and mid-sections of Baltoro Glacier is attributed to a combination of differential surface ablation resulting in debris shift between topographic highs and lows, collapse of medial moraines, and redistribution of debris following input from mass movement events, all processes that commonly occur on debris-covered glaciers (e.g. Anderson and Anderson, 2016; Hambrey et al., 1999; Hambrey et al., 2008; Heimsath and McGlynn, 2008). The presence of a sharp change in debris thickness between the main glacier tongue and Tango Glacier is attributed to variations in relative surface velocity between the two flow units and the subsequent entrainment along flow unit

boundaries. In high resolution Quickbird imagery (accessed from Google Earth (2017) on 16/01/17) a ridge at the boundary between the main glacier tongue and Trango Glacier flow units is observed, which has been mapped alongside other glaciological features such as sediment folds and ogives (Figure 10). The ridge extends from the bedrock at the up-glacier confluence between the two debris units (Figure 10a), suggesting the ridge is a medial moraine between the two flow units. Parallel to the supraglacial debris ridge are a series of deformation structures in the debris cover (Figure 10a), attributed to progressive supply and subsequent compression of debris through time as continuation of flow of the main glacier flow unit towards the terminus is constricted and blocked by the incoming flow unit of Trango Glacier. Variation in debris distribution near the terminus is further complicated by Trango Glacier displaying signs of a period of dynamic instability prior to the study period, with increasingly sinuous moraines on its surface through time (Figure 10a) and propagation of an area of high velocity along the tributary glacier's length between 2001 and 2004 (Figure 5). These geomorphological features alongside the temporal pattern observed on the glacier over the study period are consistent with a glacier that may have undergone a surge event, or at least a change in relative velocity to the ice flow unit it interacts with (Meir and Post, 1969). An arc of granitic debris that mirrors the terminus shape of Tributary Glacier 8 appears to suggest that this glacier is also dynamically linked to the terminus (Figure 10a). These geomorphological patterns suggest the main debris units were transported and deposited prior to input of debris from Tributary Glacier 8 and Trango Glacier, and indicate that initiation of debris supply along the main glacier and tributary glaciers were not contemporaneous.



551

552 **Figure 10:** A (a) geomorphological map and (b) annotated oblique Quickbird image
 553 displaying the moraine ridge structure and associated sediment folds at the boundary
 554 between Trango Glacier (Tributary glacier 9) and the main glacier tongue, and the
 555 difference in debris lithology between the two glaciers. Accessed from Google Earth (2017)
 556 on 16/01/17.

557

558 **5.2. Processes controlling debris distribution**

559 Sustained debris thickening between 2001, 2004 and 2012 was observed, although notable
 560 spatial variability exists. Sediment flux also appeared to be temporally constant across
 561 much of the glacier despite variations in surface velocity, although some small-scale
 562 variations in sediment flux did occur. Changes in sediment flux in the lower section of the
 563 glacier were considered to be a product of increasing debris thickness near the terminus
 564 and sustained surface velocity as more debris was delivered to the slow-flowing terminus
 565 area through time. Variation in sediment flux between 2001–2002 and 2003–2004 in the
 566 glacier mid-section, south of Dunge and Biale Glaciers (Tributary Glaciers 10 and 11), are
 567 attributed to a combination of increasing debris thickness and increasing area of thicker
 568 debris up-glacier of the terminus and to an increase in surface velocity, as sediment flux

569 varied considerably in this region between the two periods despite a lack of time separation.
570 However, the overall glacier-wide stability in the rate of debris thickening and the pattern of
571 sediment flux suggests that supraglacial debris transport was not the sole control of
572 spatiotemporal changes in debris layer thickening. In periods of higher velocity (e.g. 2011–
573 2012) it is likely that less debris built up on the glacier surface prior to transportation,
574 causing a thinner layer of debris to be transported down-glacier than in previous years,
575 albeit at a faster rate, and vice versa for periods of low velocity. Variability in surface
576 velocity and its influence on debris transport is particularly pertinent for Baltoro Glacier,
577 where velocity has been found to vary from year to year, observed here and by Quincey et
578 al. (2009). Longer term studies (of the order of a number of decades) considering the
579 interaction between surface velocity and debris distribution are needed to determine the
580 relationship of these two parameters over decadal to centurial timescales. Consequently,
581 the rate of debris input over sub-decadal timescales is thought to control temporal
582 variations in debris layer thickening across the glacier. Over sub-decadal timescales, debris
583 input will vary as a result of the frequency of mass movement events, which would
584 significantly increase local supraglacial debris volume and affect velocity if the volume was
585 great enough (e.g. Tovar et al., 2008). Over longer timescales (>100 years) debris input
586 would be controlled by regional erosion rates, which are in turn controlled by climatic
587 conditions, most notably precipitation, and tectonics, including rates of uplift and
588 deformation in active tectonic regions such as the Karakoram (Molnar et al., 2007; Scherler,
589 2014). Regional erosion rates therefore control the long-term (centurial to millennial) rates
590 of debris input to a glacier system, but over shorter (sub-decadal) periods the frequency
591 and location of mass movement events are important controls on spatiotemporal variations
592 in supraglacial debris distribution.

594 The total area and number of supraglacial water bodies increased between 2001 and 2012,
595 and temporal changes in these parameters were notably larger than the uncertainty
596 involved in incorrect classification of pixels containing water. The greatest percentage
597 change in supraglacial water body number (73 %) and area (171 %) occurred between
598 2001 and 2004. Increase in supraglacial water body area and number has previously been
599 attributed to changes in precipitation since 2000 (Quincey et al., 2009; Gibson et al., 2016).
600 However, increasing supraglacial water body frequency on debris-covered glaciers is often
601 considered analogous with stagnation and surface lowering of debris-covered glaciers (e.g.
602 Sakai et al., 2000). Such differential surface lowering forms the undulating debris
603 topography, which then promotes the formation of supraglacial water bodies (Hambrey et
604 al., 2008). Since 2004, Baltoro Glacier has showed no sign of stagnation but has
605 undergone surface lowering of the order of 40 m between 2000 and 2008 (Gardelle et al.,
606 2012). Such surface lowering is apparent up-glacier of the confluence between Trango
607 Glacier and the main glacier tongue, where the debris surface displays a high density of
608 topographic highs and depressions (Figure 10). Surface lowering of some glaciers in the
609 Karakoram has been attributed to negative mass balance of glaciers in response to recent
610 climatic change (Gardelle et al., 2012), although in the case of Baltoro Glacier it could
611 equally be a consequence of its tributary glaciers being in various phases of dynamic
612 instability. Glacier dynamic instability would cause temporal variation in ice flux to the main
613 glacier tongue. Following the end of these phases of dynamic instability, ice mass delivery
614 to the main glacier tongue would reduce, causing temporary reduction in surface velocity,
615 as observed between 2001 and 2004 on Trango Glacier, and thus surface lowering. To
616 understand the relative controls of climatic change and dynamic instability of tributary
617 glaciers on surface velocity and lowering of Baltoro Glacier longer-term records of surface

618 lowering, a greater record of glacier mass balance and localised meteorological data are
619 needed.

620

621 Debris thickness maps presented here show no evidence for a thicker accumulation of
622 debris at the glacier terminal margin, the presence of which has previously been interpreted
623 as a terminal moraine on maps of debris thickness for topographically confined glaciers
624 such as Khumbu Glacier in Nepal (Rounce and McKinney, 2014; Rowan et al., 2015;
625 Soncini et al., 2016). Baltoro Glacier is thought to lack such a terminal moraine due to the
626 glacier being of debris-fan-type, the occurrence of which is linked to glaciers located in
627 wide, gently sloping valleys (Kirkbride, 2000). Debris-fan termini have a steeply sloped
628 topography relative to the near horizontal glacier surface up-glacier of the terminus. The
629 presence of a sloped debris surface suggests the same is true for the underlying ice
630 surface (e.g. Figure 9), both of which would facilitate more efficient supra- and englacial
631 drainage systems and inhibit the formation of undulating topography in the supraglacial
632 debris layer near the terminus, as debris will be less stable and is more likely to be
633 transported more evenly when located on a slope. The lack of depressions near the glacier
634 terminus would therefore inhibit ponding of supraglacial water in the area.

635

636 **5.3. *Incorporating debris distribution change into numerical modelling***

637 Mean annual debris thickness change and mean annual sediment flux are potential
638 indicators to help establish the period over which a glacier has become debris covered and
639 the rate at which supraglacial debris layers evolve. Currently in numerical models of debris-
640 covered glaciers debris thickness is largely considered as static in time (e.g. Collier et al.,
641 2014; Reid and Brock, 2010; Shea et al., 2015). However, we have confirmed debris
642 distribution is dynamic over annual to decadal timescales (Figure 3; Figure 9). Incorporating

643 an annual rate of debris thickness change into long-term energy balance models for debris-
644 covered glacier surfaces is therefore important for generating robust results using these
645 methods. For glacier change models, such as those of Rowan et al. (2015), where a
646 supraglacial debris layer is formed through glacial processes and hillslope erosion rates are
647 used to control input of debris to a glacier system, annual rates of glacier change and
648 sediment flux could be used to constrain model outputs. We also confirm that using
649 temporally constant annual erosion rates for control of debris input to glacier systems, such
650 as those used by Rowan et al. (2015) and Anderson and Anderson (2016), is appropriate
651 on sub-decadal timescales, but should be set on a case by case basis as these erosion
652 rates would be affected by localised variability in headwall retreat and precipitation
653 (Bookhagen et al., 2005; Pan et al., 2010). For longer-term studies the effect of a changing
654 climate should be considered in regional erosion rates used for such numerical models
655 (Peizhen et al., 2001; Scherler, 2014). Additionally, the rate of debris layer thickness
656 change is likely to vary between glaciers due to varying input of debris, glacier size,
657 landscape, climate and bedrock lithology, and needs to be evaluated for individual cases.

658

659 To accurately determine the formation and evolution of a supraglacial debris layer a greater
660 understanding of the volume of debris contributed from englacial debris input and the role
661 varying ice velocity with depth plays in englacial debris transport is needed. At present,
662 calculation of englacial debris meltout has not been attempted in great detail (e.g. Rowan et
663 al., 2015; Anderson and Anderson, 2016). Recent work on debris-covered glaciers has
664 highlighted rockfall in accumulation areas can be incorporated rapidly to englacial locations
665 (Dunning et al., 2015), but very little is known regarding the volume of debris contained
666 within the glacier ice of debris-covered glaciers (Anderson 2000). Enhanced ablation and
667 surface lowering, as seen on Baltoro Glacier from the start of the 21st century (Gardelle et

668 al., 2012) is likely to result in an increased rate of debris meltout (Bolch et al., 2008;
669 Kirkbride and Deline, 2013). By quantifying the volume of debris contributed to a glacier
670 surface through englacial meltout a more comprehensive understanding of processes by
671 which debris distribution is controlled, both through space and time, could be gained. Such
672 data have previously been collected through the use of ground penetrating radar (e.g.
673 McCarthy et al., 2017), but a greater spatial coverage of such data across glacier surfaces
674 is needed to understand spatial variability in englacial debris distribution.

675

676 **6. Conclusion**

677 The distribution of supraglacial debris on Baltoro Glacier predominantly follows the
678 expected pattern for a debris-covered glacier, with increasingly thick debris towards the
679 terminus. However, debris distribution is complicated by the interaction between tributary
680 glaciers, some of which show signs of dynamic instability, and the main glacier tongue. An
681 overall increase in debris thickness was observed between 2001 and 2012, indicating that
682 supraglacial debris distribution varies over sub-decadal timescales. Short-term variations in
683 debris thickness are primarily attributed to input from mass movement events. The area of
684 Baltoro Glacier covered by a spatially continuous debris layer increased over the study
685 period, suggesting that the debris layer is still evolving. The number and area of
686 supraglacial water bodies on Baltoro Glacier also increased through the study period, with
687 changes attributed to differential surface lowering. However, ponding is not observed at the
688 terminus because the glacier displays a debris-fan type terminus that inhibits formation of
689 undulating debris topography and facilitates efficient drainage. Additionally, surface
690 lowering of the glacier surface up-glacier of the terminus may be important for debris layer
691 thickening due to exhumation of debris transported englacially.

692

693 Quantifying the influence of mass movement deposits and englacial meltout on supraglacial
694 debris distribution is important to better understand the evolution of debris-covered glaciers
695 through time, particularly to determine the mass balance of glaciers accurately in response
696 to recent and future climatic change. However, quantifying such inputs is challenging;
697 mass movement events are temporally and spatially variable and dependant on climate,
698 topography, tectonic processes and lithology, and identifying debris contributed from
699 englacial sources requires quantification of the volume of debris held englacially, which can
700 only really be gained through fieldwork. Despite such limitations, this study shows that
701 incorporating some aspects of spatiotemporal change in supraglacial debris distribution into
702 numerical modelling is achievable, and is likely to be significant in accurately determining
703 debris-covered glacier systems.

704

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708 an anonymous reviewer for their thorough and constructive reviews of the manuscript.

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718 **Author contributions**

719 Morgan Gibson – conceived the study, carried out the data processing and analysis and
720 wrote the manuscript.

721 Neil Glasser – contributed to data analysis and refinement of discussion

722 Duncan Quincey – contributed to data analysis and final editing of the manuscript

723 Ann Rowan – contributed to idea development and final editing of the manuscript

724 Christoph Mayer – collection of field data, contributed to data analysis and refinement of
725 discussion

726 Tristram Irvine-Fynn – contribution to wording in the introduction and discussion of the
727 manuscript, and editing of the manuscript.

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